Lithospheric velocity structure of the New Madrid Seismic Zone: A joint teleseismic and local P tomographic study

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[1] The enigmatic seismicity in the New Madrid Seismic Zone (NMSZ) has been attributed to some abnormal lithospheric structure, including the presence of dense mafic intrusions and a low-viscosity lower crust. However, the area’s detailed lithospheric structure remains unclear. Here we invert 2,056 teleseismic P and 12,226 local P first arrival times from a recent nine-year dataset to infer the lithospheric velocity structure beneath the NMSZ. Our results show that the seismically active zone is associated with a local, NE–SW trending low-velocity anomaly in the lower crust and upper mantle, instead of high-velocity intrusive bodies proposed in previous studies. The low-velocity anomaly is on the edge of a high-velocity lithospheric block, consistent with the notion of stress concentration near rheological boundaries. This lithospheric weak zone may shift stress to the upper crust when loaded, thus leading to repeated shallow earthquakes.


1. Introduction

[2] The New Madrid Seismic Zone (NMSZ), located in the northern Mississippi embayment, is the most seismically active region in the Central and Eastern United States (CEUS) (Figure 1). A sequence of at least three major earthquakes (Mw ≥ 7.0) occurred here during the winter of 1811–1812 [Johnston and Schweig, 1996], and thousands of microearthquakes have been recorded since 1974. The microseismicity delineates three linear faults in the NMSZ (Figure 1): (1) the NE-trending Blytheville Fault Zone (BFZ), (2) the NW-trending Reelfoot Fault (RF), and (3) the NNE-trending New Madrid North Fault (NN) [Johnston and Schweig, 1996]. The largest events including the 1811–1812 main shocks [Johnston and Schweig, 1996; Hough et al., 2003], the 1843 Marked Tree, Arkansas earthquake (M ~ 6.3), and the 1895 Charleston, Missouri earthquake (M ~ 6.6) [Johnston, 1996] are thought to have occurred on those faults (Figure 1). Paleoseismologic studies also suggest that several large earthquakes similar to the 1811–1812 sequence have happened in the NMSZ in the past a few thousand years [Tuttle et al., 2002].

[3] The cause of these intraplate earthquakes remains uncertain. The NMSZ is located within the Reelfoot rift [Ervin and McGinnis, 1975] which may be related to the seismicity [Johnston and Kanter, 1990], but not all rifts in the CEUS are seismic [Li et al., 2007]. The surface deformation associated with the NMSZ is minimal, and recent GPS studies, while differing in details, show near zero site velocities outside the NMSZ [Newman et al., 1999; Smalley et al., 2005; Calais et al., 2005; Calais and Stein, 2009]. Hence some “deep” and “local” causes have been suggested to explain the seismicity. One such cause is a recent change to the dense Proterozoic-Cambrian mafic intrusions that provides a localized gravitational force in the rift [Grana and Richardson, 1996; Pollitz et al., 2001]; another is low-viscosity lower crust (and probably low-viscosity upper mantle too) under the NMSZ that can shift deviatoric stresses imposed by tectonic (stress or thermal) perturbations to the upper crust to trigger a sequence of earthquakes [Kenner and Segall, 2000]. However, the details of lithospheric structure under the NMSZ are not clear. Mitchell and co-workers found some low-velocity anomalies in the NMSZ lithosphere [Mitchell et al., 1977; Al-Shukri and Mitchell, 1987], but the resolution of their models is limited due to the data available at that time.

[4] In this study we use a new dataset recorded between 1999 and 2007 from the Cooperative New Madrid Seismic Network (Figure 1) operated by the Center for Earthquake Research and Information (CERI) to map the lithospheric structure beneath the NMSZ. The network contains three times the number of seismic stations in a much more focused study area compared to Al-Shukri and Mitchell’s [1987].

2. Data and Models

[5] From this new dataset, we extracted both teleseismic and local P first arrivals for the joint tomographic inversion. We picked 121 teleseismic events (Figure S1 of the auxiliary material) using the criteria of (1) magnitude Mb ≥ 5.0, (2) epicenter distance ≥ 25°, and (3) recording station number ≥ 8.1 Theoretical first-arrival times of teleseismic P phases were calculated from the AK135 model [Kennett et al., 1995], and then relative travel-time residuals were measured for each event by a semi-automated method, Multi-Channel Cross-Correlation (MCCC) [VanDecar and Crosson, 1990]. We modified the method by running the MCCC multiple times with shift corrections for all correlation windows after each run. When the shift corrections converge to zero, the correlation windows used for calculating relative delays are adjusted to the right corresponding positions. This procedure improves the measurements by up to 0.5 sec. During the process, we used a threshold
value (0.06 sec) of root-mean-square timing uncertainty [VanDecar and Crosson, 1990] to rule out the low-quality waveforms. A total of 2,056 accurate teleseismic P travel-times were selected for this study. 95% of the teleseismic residuals relative to the network mean vary within a range of ±0.5 sec.

[5] The poor vertical resolution of teleseismic tomography can be improved by including local P phases to constrain the shallow velocity structure and separate the residual contributions between the crust and mantle. We extracted local P picks from the CERI catalog with the following requirements: (1) pick accuracy within 0.35 sec, (2) a minimum of 8 recording stations for each event, and (3) a minimum of 8 event records for each station. Those criteria yielded 12,226 joint first arrival measurements of teleseismic and local P phases for our inversion.

[7] These travel-time data were corrected for both station elevation and alluvial sediment thickness in the NMSZ. The latter is necessary because the unconsolidated sediment layer has a very low P wave velocity (1.8 km/sec) [Chiu et al., 1992], and its thickness varies significantly (0–1000 m) in our study area (Figure 1), causing a large travel-time variation (up to ~0.5 sec). To correct the effects of the sedimentary cover, we used two independent methods, hereinafter referred to as M1 and M2 (refer to Figure S2 of the auxiliary material for the correction values). Method M1 calculates the correction terms directly by taking Bodin et al.’s [2001] sediment thickness measurements (Figure 1) based on hundreds of well logs [Dart, 1992], and assuming vertical ray paths within the sediment layer. The weakness of this method is that the well-data shortage in some locations may cause biased correction values, especially for where the thickness changes substantially. Method M2 follows Vlahovic et al. [2000] and Vlahovic and Powell’s [2001] idea that the combined correction terms of station elevation and sediment thickness can be replaced by the average station residuals of local phases. This method requires similar sampling ray paths for each station, which in reality is not strictly satisfied. Thus neither method is perfect. Nonetheless, the consistent tomographic results after the two unrelated corrections can be considered “immune” to the sediment effects.

[8] The 1-D initial velocity model for our joint tomographic inversion is listed in Table 1. The crust (≤40 km) consists of three layers simplified from a crustal model of Chiu et al.’s [1992]. The upper mantle (40–160 km) comprises two layers whose velocity values are based on a recent refraction profile [Catchings, 1999]. The cell size for our tomography has a fixed horizontal scale of 15 km × 15 km and a variable vertical scale depending on the initial velocity layers. Slowness in each cell was resolved iteratively by using a nonlinear 3-D tomography method. During the inversion, Laplacian damping was used to balance the resolution and smoothness of the velocity models by trial-and-error. After the inversion, our tomographic models achieved 43.7% and 30.8% variance reductions of travel-time errors from the starting model for the M1 and M2 methods, respectively.

### Table 1. 1-D Initial Velocity Model for Joint Tomographic Inversion

<table>
<thead>
<tr>
<th>Layers, km</th>
<th>P velocity, km/s</th>
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<tr>
<td>0–5</td>
<td>4.22</td>
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<tr>
<td>5–17</td>
<td>6.17</td>
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<tr>
<td>17–40</td>
<td>6.98</td>
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<tr>
<td>40–100</td>
<td>8.3</td>
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<td>100–160</td>
<td>8.4</td>
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0.1 sec standard deviation (the root-mean-square of the residuals after the true inversion) to those synthetic travel-times. The inverted tomographic result from the synthetic data is shown in Figure 2. Although some smearing occurred in the NE–SW direction due to the uneven distributions of the events and stations, the alternating patterns were well recovered for most areas. Moreover, a separate sediment leaking test demonstrates that the low-velocity top layer does smear down for our inversion (Figure S3 of the auxiliary material).

Figures 3a and 3b show the P velocity structures beneath the NMSZ resulting from inverting travel-times with the M1 and M2 corrections, respectively. The major difference is within the upper crust (0–5 and 5–17 km layers), where correction M1 yields a strong low-velocity anomaly to the north of the Reelfoot rift (Figure 3). This could be an artifact introduced by the lack of drill-hole data [Dart, 1992]. Furthermore, the anomaly is shallow and beyond the area of our primary interest. Other than for the top two layers, the M1 and M2 corrections produce almost identical velocity images (17–160 km). Therefore the deep structures are not significantly affected by the sediment using our joint dataset. Our analyses will focus on the features common in Figures 3a and 3b.

[11] The prime common feature of Figures 3a and 3b is the low-velocity (up to −3%) zone in the lower crust (17–40 km) and upper mantle (40–160 km). It is spatially associated with three seismic segments in the NMSZ, although its locations vary somewhat from layer to layer. Within the lower crust (17–40 km), the low-velocity anomaly is distributed along the BFZ; for the 40–100 km layer, the low-velocity anomaly is concentrated around the RF and southwestern tip of the BFZ; for the 100–160 km layer, the low-velocity anomaly broadens and covers much of the southwestern Reelfoot rift. The low-velocity zone is surrounded by scattered high-velocity (up to 3%) blocks and primarily confined within the Reelfoot rift, with an overall trend parallel to the rift axis. It also shows a strong spatial relationship with both historic large events and microseismicity.

4. Discussion

[12] Our tomographic results are generally consistent with those from previous studies. Vlahovic et al. [2000] used local events to construct the P wave velocity structure in the NMSZ for the top 10.65 km. Their top two layers (<2.65 km) and underlying layers (2.65–10.65 km) show respectively high-velocity and somewhat low-velocity anomalies in the seismic zone, similar to our results. The high-velocity anomaly in the top layer (0–5 km) in our models is consistent with the stable surface indicated by the GPS data [Newman et al., 1999; Calais et al., 2005; Calais and Stein, 2009].

[13] The low-velocity structures we image beneath the Reelfoot rift are similar to those in previous studies using teleseismic P data alone [Mitchell et al., 1977; Al-Shukri and Mitchell, 1987]. When incorporating local P data, Al-Shukri and Mitchell’s [1987] model shows a belt of slightly higher velocity (up to 1%) in the lower crust of the seismic zone. But the belt is primarily interpolated from the high-velocity data blocks on their model margins. Furthermore, because almost all local events in the NMSZ are too shallow to sample the lower crust, a sufficiently large dataset is needed to resolve the vertical trade-off of delay times. Previous refraction seismic profiles [Mooney et al., 1983; Catchings, 1999] suggest high-velocity intrusions at the bottom of the lower crust, but their interpreted intrusive bodies are broader than the rift zone or the scale of our lower crust image.
[14] Teleseismic tomography inverts relative travel-time residuals and thus only shows relative velocity variations. Regional tomography studies can help constrain the absolute velocity of the study area and understand the context of the NMSZ within the stable North American craton, although their large cell sizes and smoothing may make them unable to resolve small-scale anomalies such as those in our study. Regional surface wave and Pn tomography models [Van der Lee and Frederiksen, 2005; Zhang et al., 2009] both indicate that the NMSZ is on the edge of a high-velocity block in the upper mantle, with slightly higher absolute velocity (Figure S4 of the auxiliary material). Liang and Langston's [2008] ambient noise tomography in the CEUS also shows that the NMSZ is on the western boundary of a high-velocity block for 15 sec period which samples almost the whole crust. The results suggest that the

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**Figure 3.** (a) Lithospheric P velocity structure for the NMSZ using sediment correction method M1. First P arrivals of 2,056 teleseismic phases and 12,226 local phases were used for the joint inversion. The notions of thick grey lines, stars and diamonds follow Figure 1. (b) Counterpart of Figure 2a using sediment correction method M2. Both methods show a low-velocity zone in the lower crust and upper mantle, beneath the seismicity.
low-velocity anomaly in the NMSZ is a localized feature on the rim of a rigid cratonic root.

[15] The cause of the low-velocity zone in the NMSZ lithosphere is uncertain. The anomaly is localized within the Reelfoot rift and generally follows its trend, and therefore may be genetically linked to the rift. On the other hand, given that the Reelfoot rift initiated in late Precambrian [Ervin and McGinnis, 1975] and the NMSZ lacks clear thermal anomaly [McKenna et al., 2007], the low-velocity anomaly is unlikely to have a thermal origin. Alternatively the low-velocity zone might represent compositional variations across the rift, perhaps related to the magmatism during the Reelfoot rifting. But our results do not favor intrusive bodies in the lower crust because mafic intrusions typically cause high-velocity anomalies within rifts [Mooney et al., 1983]. These lead us to speculate that the anomaly is related to the deformational fabrics (weak zone). Although most recent GPS data show little strain rate in the NMSZ and its surrounding area [Newman et al., 1999; Calais et al., 2005; Calais and Stein, 2009], the structural and geometrical analyses suggest that the slip rates of the NMSZ faults could be as high as 4.4–6.2 mm/yr over the last a few thousand years [Mueller et al., 1999; Van Arsdale, 2000]. Alternatively, this low-velocity feature could be a preserved weak zone in the lithosphere from the last major tectonic event (e.g., Paleozoic or Mesozoic) within the Reelfoot rift.

[16] The locations of the major intraplate seismic zones in the CEUS (the NMSZ, the East Tennessee Seismic Zone, the Charleston Seismic Zone, and the New England Seismic Zone) are spatially correlated with high-velocity block edges or velocity transition zones in the lithosphere [Liang and Langston, 2008; Zhang et al., 2009]. Similar scenarios can be found in other significant intraplate seismic zones such as the Shanxi rift, China [Tian et al., 2009] and the Kutch rift, India [Kennett and Widiyanto, 1999] which are underlain by a low-velocity zone (presumably a weak zone) bordering a rigid lithospheric root. Those structures suggest that the rheological contrast may play an important role for the intraplate earthquakes. Numerical modeling shows that rheological boundaries tend to localize stress near thin lithosphere [Li et al., 2007], and a localized weak zone in the lower crust is likely to shift stress to the upper crust by viscous relaxation. If our lithospheric low-velocity zone in the NMSZ reflects weakness, Kenner and Segall [2000] suggest that such a stress-shifting process can lead to repeated shallow earthquakes. Although present geodetic observations do not detect enough surface motion or strain for large earthquakes [Newman et al., 1999; Calais et al., 2005; Calais and Stein, 2009], it is possible that strain accumulation remains slow during interseismic cycles, especially after large energy is released.

5. Conclusions

[17] We have modeled the lithospheric velocity structure of the NMSZ using a combination of teleseismic P and local P travel-time data. Two independent methods correcting for sediment effects yield almost identical velocity variations below the upper crust. Our tomographic results show that the NMSZ faults are underlain by a localized low-velocity anomaly in the lower crust and upper mantle which is primarily confined within and parallel to the Reelfoot rift. On the other hand, our results do not show compelling evidence for dense mafic intrusions in the lower crust that have been proposed. The low-velocity anomaly under the NMSZ may represent a deep shear zone at rheological boundaries. Such a weak zone could shift stress to the upper crust, thus help explain the repeated earthquakes in the NMSZ where the present-day strain rate is near zero.

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